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Why is Africa rifting?

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Abstract: Continental rifting has a fundamental role in the tectonic behaviour of the Earth, shaping the surface we live on. Although there is not yet a consensus about the dominant mechanism for rifting, there is a general agreement that the stresses required to rift the continental lithosphere are not readily available. Here we use a global finite element model of the lithosphere to calculate the stresses acting on Africa. We consider the stresses induced by mantle flow, crustal structure and topography in two types of models: one in which flow is exclusively driven by the subducting slabs and one in which it is derived from a shear wave tomographic model. The latter predicts much larger stresses and a more realistic dynamic topography. It is therefore clear that the mantle structure beneath Africa plays a key part in providing the radial and horizontal tractions, dynamic topography and gravitational potential energy necessary for rifting. Nevertheless, the total available stress (c. 100 MPa) is much less than that needed to break thick, cold continental lithosphere. Instead, we appeal to a model of magma-assisted rifting along pre-existing weaknesses, where the strain is localized in a narrow axial region and the strength of the plate is reduced significantly. Mounting geological and geophysical observations support such a model.



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Rifts play an important part in shaping our planet, breaking continents apart and eventually leading to the formation of ocean basins. The topography they generate affects our climate and gives rise locally to fertile regions, which have had an important role in human evolution (Bailey *et al.* 2000; King & Bailey 2006). The East African Rift (EAR) is the best example of a continental rift currently active on Earth; it captures the early stages of rift development in southern Africa through to incipient oceanic spreading in Afar. This rift environment hosted the earliest hominoid evolution and is the site of numerous natural resources, including freshwater, minerals containing the rare earth elements and precious metals. It is also a key source of energy for Africa in the form of geothermal, hydroelectric and petroleum reserves.

Although continental rifting plays a fundamental part in plate tectonics, the driving forces behind rifting are not readily apparent. In the so-called plate stretching models, distant forces are invoked to facilitate rifting, usually in the presence of a mantle plume, which helps to thermally weaken the base of the plate (White & McKenzie 1989). Mantle hotspots lead to uplift and, in a series of papers, Burke and Dewey argued for the development of triple junctions above these upwellings (e.g. Burke

& Dewey 1973). With time, the arms of a triple junction compete for dominance. Burke and Dewey argued that multiple upwellings beneath Africa eventually linked up to form the EAR and the Cameroon volcanic line. Pre-existing weaknesses – along suture zones or sites of previous rifting, for example – may control the site of rifting (Vauchez *et al.* 1997; Corti *et al.* 2007). The strength and morphology of the plate will also play a part in shaping the geometry of the rift (Ebinger & Sleep 1998; Ebinger *et al.* 1999; Ebinger 2005).

A number of researchers have argued that the rifting forces available are insufficient to rupture thick, cold lithosphere (e.g. Kuszniir & Park 1987), leading to a plate force paradox (Karner *et al.* 2004). However, calculating the forces available is not a trivial exercise. There is a tendency to treat the plate-driving mechanisms in isolation and as separate from mantle convection when estimating the stresses available – for example, slab forces v. gravitational potential energy (GPE). The relative contribution from various mechanisms will vary with the locality and any estimates must consider the process as a coupled system on a global scale (Hager & O'Connell 1979; Bai *et al.* 1992).

To overcome the plate force paradox, melt can be invoked as a mechanism that weakens plates

and localizes the accommodation of strain to magmatic segments (e.g. Buck 2004). There is now abundant evidence for focused melt distribution throughout the uppermost mantle and crust in the EAR (e.g. Keir *et al.* 2011; Desissa *et al.* 2013; Hammond *et al.* 2013; Stork *et al.* 2013; Hammond & Kendall 2016). Based on laboratory experiments and seismic observations, Holtzman & Kendall (2010) argued for melt partitioning and strain localization as a mechanism that lubricates plate boundaries and facilitates rifting in both continental and oceanic settings.

This paper first reviews the driving forces for rifting and then the factors that control the strength of a tectonic plate. We then conduct an inventory of the forces available for rifting in Africa, drawing on the work of Lithgow-Bertelloni & Gynn (2004). As in Lithgow-Bertelloni & Gynn (2004), we take a global dynamic perspective, treating the plates and mantle as a coupled dynamic system (Lithgow-Bertelloni 2014). In this view, the so-called distant plate forces, such as slab-pull, are the result of the negative buoyancy of the slabs, which drives convection and generates traction at the base of all the plates. We also account for the contributions of GPE associated with lithospheric structure, including both the isostatic and dynamic topography. We then summarize the geophysical evidence for melt distribution in the lithosphere of the EAR system. We conclude with a summary of the forces available, a discussion of the rifting mechanisms in Africa based on observations, and some of the remaining challenges in fully understanding the mechanisms that lead to rifting.

Driving forces of rifting

What drives plate motions is an enduring question that has been framed as plates *v.* mantle. An emphasis has historically been placed on considering the stresses at plate boundaries, where the far-field extensional forces are primarily attributed to plate subduction (Forsyth & Uyeda 1975). This is akin to pulling a tablecloth across a table by pulling on one end. An obvious challenge with explaining rifting in Africa in this manner is that the Nubian and Somalian plates are surrounded by mid-ocean ridges and it is difficult to invoke distant extensional forces as the driving mechanism for rifting. Instead, studies of mantle convection have shown that the plates are the uppermost thermal boundary layer that moves with the convecting system (Turcotte & Oxburgh 1967) – a bit like viewing our tablecloth as being transported on a conveyor belt.

Because plates are the upper thermal boundary layer of the convecting mantle system, viewing plate-driving forces as an inventory of forces acting

at plate boundaries misses the fundamental nature of the plate–mantle system: plates are mantle convection (e.g. Bercovici 2003) and the only driving force is gravity acting on lateral variations in density and the only resisting forces are due to the strength of the material, e.g. the mantle and crust. The difficulty for the Earth arises largely from the latter. We have a pretty good idea of the sources of buoyancy generating lateral heterogeneities in density, but our knowledge of the complex rheology of rocks as a function of pressure, temperature, composition and deformation (stress and strain rate) is limited. Cataloguing plate-driving forces in terms of boundary forces is a way of parameterizing our uncertain knowledge of the rheology of the mantle, the lithosphere and, in particular, the plate boundaries (Lithgow-Bertelloni 2014).

Treating the system as one with varying degrees of rheological sophistication reproduces plate motions well at the global level (Lithgow-Bertelloni & Richards 1998; Becker & Faccenna 2009; Gosh & Holt 2012; van Summeren *et al.* 2012) and even at the regional level of smaller plates or regions where the plate boundary rheology becomes significant (Stadler *et al.* 2010).

From this perspective, we can still speak of ridge-push and slab-pull by bookkeeping of the appropriate buoyancy sources. Ridge-push results from the thickening and densification of the oceanic lithosphere as it ages. Slab-pull results from the buoyancy of the slab. Gravity acting on this density heterogeneity gives rise to flow over the entire mantle, generating tractions that act at the base of all plates. The viscous stresses that resist the flow may be modified by the rheology. Strong slabs might act as stress guides, supported by their own weight, transmitting the generated stresses straight to the plate and inducing very little flow (Conrad & Lithgow-Bertelloni 2002). Hence all buoyancy sources generate driving traction, which is resisted by viscous or elastic stress.

Additional sources of buoyancy (from the bottom thermal boundary layer and/or internal heating, or the density structure of the plates, such as continent–ocean density differences and pressure gradients due to isostatic or dynamic topography) can generate smaller contributions to the tractions and hence the global balance (<10% of the total), although they may be significant at the regional level (Lithgow-Bertelloni & Silver 1998; Steiner & Conrad 2007). One way to account for the contribution from the lithospheric structure and topography is to calculate the equivalent GPE and its potential effects on the plate-driving forces and stresses (Frank 1972; Artyushkov 1973; Gosh & Holt 2012).

We considered the forces available on the African plate from this perspective. These include:

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buoyancy forces in the mantle, which give rise to horizontal tractions that drive plate motions and radial tractions that generate the dynamic topography; resisting tractions due to the viscosity structure of the plate and mantle; and the GPE due to the structure of the crust and lithosphere.

In the context of Africa, each of these mechanisms has been discussed previously, but often in isolation. For example, Stamps *et al.* (2010) considered the role of GPE as a driving force for rifting. In contrast, Quéré & Forte (2006) argued that mantle tractions alone are sufficient to drive rifting in East Africa. In reality, no mechanism works in isolation; here, we attempt to quantify the contributions for Africa within a global model.

In terms of rifting, however, it is not enough to have an inventory (at times incomplete) of the tractions driving plates. It is also necessary to consider how these tractions drive deformation in an elastic or viscoelastic lithosphere. In other words, the stresses driving rifting or any deformation are mediated by the response of the shell on which they act. For example, horizontal tractions of the order of 10 MPa may give rise to very large resultant stresses in an elastic lithosphere (Lithgow-Bertelloni & Guynn 2004). The tectonic stress field is controlled by sources both within the plate and those associated with mantle flow over long timescales and over a broad range of length scales.

Here we draw on the results from a finite element model of the lithosphere that was used to calculate the stresses induced by mantle flow, crustal structure and topography (Lithgow-Bertelloni & Guynn 2004; Naliboff *et al.* 2009, 2012). The focus is on longer wavelength processes (*c.* 200 km) beyond the flexural wavelength. For ease of comprehension, we separate the sources of stress that originate from buoyancy in the mantle below the lithosphere (in the form of horizontal (shear) and radial (normal) tractions) and sources of stress that are due to intra-lithospheric heterogeneity. Lithgow-Bertelloni & Guynn (2004) considered mantle flow derived from a model in which mantle flow was exclusively driven by slab subduction (the slab model) (Lithgow-Bertelloni & Richards 1998) and shear wave tomographic models (Grand *et al.* 1997; Ritsema *et al.* 1999). The slab model is derived by dropping viscous slabs into the mantle at the trench for 12 tectonic stages spanning the time range from 150 Ma to the present. The slabs sink vertically in the upper mantle at a terminal velocity equal to the convergence velocity at the trench. An effective dip angle is achieved due to trench migration between tectonic stages. The sinking velocity slows when the slabs enter the lower mantle by a factor (of about four) proportional to the viscosity contrast between the upper and lower mantle. Each slab has a density contrast with respect to the

ambient mantle that is proportional to its age at the time of subduction. The resulting three-dimensional mass distribution agrees well to a first order with tomographic images and gives rise to excellent fits to the plate motions and the gravity field. For the tomographic models, shear wave anomalies were converted to density by a constant conversion factor of $0.2 \text{ g cm}^{-3}/\text{km s}^{-1}$. Lithgow-Bertelloni & Guynn (2004) further considered two models of lithospheric heterogeneity based on the crustal model CRUST 2.0 (Bassin *et al.* 2000): one of these models enforced isostatic compensation, whereas the other did not. We focused on the latter, because we feel that it is more realistic in the context of Africa.

Lithgow-Bertelloni & Guynn (2004) calculated both the horizontal and radial tractions that arise from the three-dimensional distribution of buoyancy in the mantle from either the slab model or tomography by solving the equations for the conservation of mass and momentum in a spherical shell for a Newtonian fluid, where the viscosity increases only with depth. The equations were solved analytically via propagator matrices (Hager & O'Connell 1979) using spherical harmonic expansions to harmonic degree and order 50 when appropriate. The radial profile of viscosity includes a lithosphere moderately more viscous than the sub-asthenospheric part of the upper mantle (by a factor of 10), a low viscosity asthenosphere (factor of 0.01) and a strong lower mantle (factor of 50). This viscosity profile gives an excellent match to the present day geoid anomalies for the slab model (Lithgow-Bertelloni & Richards 1998). For consistency, the same viscosity profile was used for all buoyancy fields. The GPE fields were calculated from CRUST 2.0 assuming a 100 km depth of compensation and a uniform lithospheric mantle.

The intra-lithospheric stresses were calculated using the finite element model described in technical detail in Lithgow-Bertelloni & Guynn (2004). In brief, each source of stress was applied separately to a spherical shell of resolution $2 \times 2^\circ$ and 100 km in thickness. The stress resultants were calculated by solving the momentum equation for an elastic solid using the finite element package ABAQUS. As the rheology was linear, the results can be superposed. Horizontal tractions were applied at the bottom of the spherical shell. In the absence of radial tractions and bending stresses, horizontal tractions generate almost purely strike-slip stresses as per Love's thin shell equations (Lithgow-Bertelloni & Guynn 2004).

The contributions due to radial tractions (dynamic topography) cannot be applied directly as they create very large bending stresses in the absence of the gravitational restoring force. Instead, we separately calculated the contribution due to the elevation (gravity stresses) and those due to

extension and compression of the shell (membrane stresses); see also Lithgow-Bertelloni & Guynn (2004, fig. 2).

The stresses resulting from the lithospheric structure (GPE) were calculated by applying the pressure obtained from the calculation of the GPE to a depth of compensation of 100 km to each of the lateral faces of the elements in the shell, as in Richardson & Reding (1991).

The in-plane principal stresses (i.e. assuming that the maximum stress was in the vertical direction) were extracted from the full three-dimensional stress field calculated. The implied regime (normal, compressional and strike-slip) was calculated using the formalism of Simpson (1997).

Strength of the continental lithosphere

A number of factors influence the strength of a plate. These include the composition, age, temperature, style of deformation and pore pressure (e.g. Brace & Kohlstedt 1980). The continental lithosphere consists of a quartz-rich crust that overlies an olivine-rich mantle. This leads to a rheologically layered plate, which is weakest in the lower crust, where quartz deforms plastically. This model is often referred to as the 'jelly sandwich' model (e.g. Jackson 2002), where the mantle is the strongest part of the lithosphere. In contrast, oceanic lithosphere is stronger because it lacks a thick quartz-rich crust, which can explain why rifting is more common in the continental lithosphere (Vink *et al.* 1984). This model is not universally accepted and arguments have been made that the strength of the lithosphere is controlled by a single seismogenic crustal layer (Maggi *et al.* 2000; Jackson 2002). The strength envelopes will be strongly affected by the amount of water in the lower crust and upper mantle. A wet uppermost mantle may be much weaker than is commonly thought.

Mechanical stretching of the lithosphere as a response to distant forces can be thought of as an end-member style of rifting (e.g. McKenzie 1978; Wernicke 1985). In such models, extension is accommodated by large offset border faults in the brittle parts of the crust and a broad zone of stretching in the more ductile parts of the plate. These models predict the formation of sedimentary basins and many features seen at passive margins (e.g. McKenzie 1978). However, a problem with these models is that they require very large stresses to rupture the lithosphere – hence the 'plate force paradox' (Buck 2004). The yield strength will vary with factors such as the strain rate, flow parameters, the water content and heat flow. Stamps *et al.* (2010) showed how the yield strength of the East African lithosphere can increase by a factor of two on

moving from north to south. In general, the predicted yield stresses vary from 200 MPa to 1 GPa depending on the assumed parameters.

The mechanical or tectonic stretching models neglect the effects of magmatism and associated heating in the rifting process. Figure 1, from Buck (2004), illustrates the features of the tectonic stretching model (in a 'jelly sandwich' plate) and a model of magma-assisted rifting. Both models assume a 30 km thick crust with a strain rate of 10^{-14} s^{-1} and a thermal profile derived from a surface heat flow of 40 mW m^{-2} ; see Buck (2004) for details. In the magma-assisted model, the strain is localized to the rift axis, predicting much sharper lateral changes in crustal thickness and the depth to the lithosphere–asthenosphere boundary. The yield strengths for these models are much smaller than those for the tectonic stretching models.

Pre-existing weaknesses and lithospheric thin spots

The seminal paper of Wilson (1966) proposed the idea that ocean basins open and close in a cyclical nature; it has long been recognized that pre-existing weaknesses in the crust can serve as nucleating points for rifting. In fact, Dunbar & Sawyer (1989) argued that such features could explain the differences between magmatic and amagmatic rifting. However, we now understand that this difference more probably reflects different stages of rifting, where the strain is increasingly accommodated by magmatism as a rift system matures (Ebinger & Casey 2001).

Rifting may be influenced by the rock fabric and the style of deformation. Vauchez *et al.* (1997) discussed the frequently observed correlation between rifts and the pre-existing tectonic fabric (Fig. 2). They argued that an inherited mechanical anisotropy in the lithosphere may preferentially weaken the lithosphere in directions parallel to orogenic belts. Rift-parallel seismic anisotropy may provide the signature of this fabric, but such observations have also been attributed to melt-induced anisotropy (e.g. Kendall *et al.* 2005, 2006; Holtzman & Kendall 2010). Nevertheless, both mechanisms serve to weaken the lithosphere. In Africa, the western branch of the Tanzanian portion of the EAR is thought to follow the weaker lithosphere of the mobile belts (Nyblade & Brazier 2002) and the Main Ethiopian Rift is thought to be the site of a Proterozoic collision zone (Keranen & Klemperer 2008; Cornwell *et al.* 2010). Elevated temperatures during rifting may lead to a reduction in grain size that facilitates a transition from dislocation to diffusion creep mechanisms, further reducing the strength of rift boundaries (Hopper & Buck 1993).

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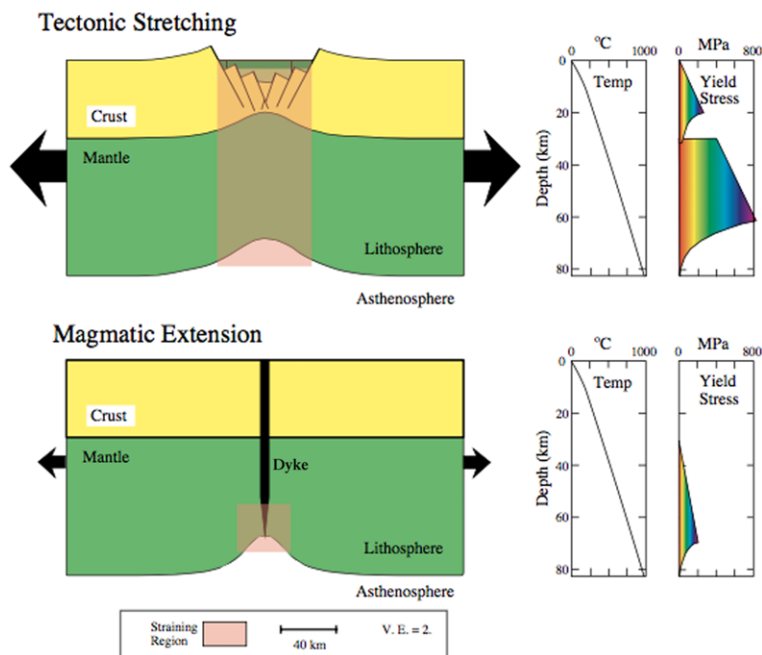


Fig. 1. Comparison of a mechanical or tectonic stretching model (upper panel) and a magma-assisted rifting model (lower panel) for a given geotherm and strain rate (redrawn from Buck 2004). The yield stress as a function of depth is shown on the far right and is much smaller for the magma-assisted model.

It is well known in rock mechanics that pre-existing weaknesses play a central part in deformation. Figure 2b, derived from Hall *et al.* (2006), shows a laboratory-scale example of a plate with two pre-existing weaknesses that had been subjected to compressive stress. As the plate ruptured, fractures propagated away from the pre-existing weaknesses. Interestingly, the 'rift' in the plate bifurcates, resembling the western and eastern arms of the EAR. Although this is a very qualitative comparison, it demonstrates the role that pre-existing weaknesses can have in shaping the geometry of rift systems. Lateral variations in the strength of the plate will shape the dimensions of the rift valley and flank uplift (Ebinger *et al.* 1999).

Lithospheric thin spots may also be required to initiate rifting. Armitage *et al.* (2009) showed how a thin lithosphere is required for a thermal anomaly to initiate melt production by decompression. Ebinger & Sleep (1998) argued that a single plume interacting with a thin lithosphere beneath the Mesozoic–Palaeogene rifts and passive margins of Africa and Arabia may guide the lateral flow of plume material. The structure of the lithosphere appears to exert a strong influence on the spatial distribution of plume-related melting and magmatism.

Both pre-existing weaknesses and lithospheric thin spots can help to produce the decompression melting that facilitates rifting. The melt serves to focus the strain with little crustal stretching and can significantly reduce the strength of plate (Buck 2004).

Dynamic topography and the African super-plume

Africa, as a continent, sits anomalously high in elevation. Whereas most cratons have an average elevation of a few hundred metres, large portions of Africa stand more than 1 km above sea-level (Lithgow-Bertelloni & Silver 1998). It is difficult to explain this topography by isostatic compensation due to thickness and density variations, given the known crustal and lithospheric structure of Africa. Instead, it is now commonly accepted that mantle flow driven by density anomalies in the mantle acts to dynamically support the excess elevation in Africa (Lithgow-Bertelloni & Silver 1998) and may even explain the uplift of the western margin of the Arabian craton (Daradich *et al.* 2003).

Africa is underlain by a large low shear velocity province (LLSVP) (Fig. 3), which is one of two

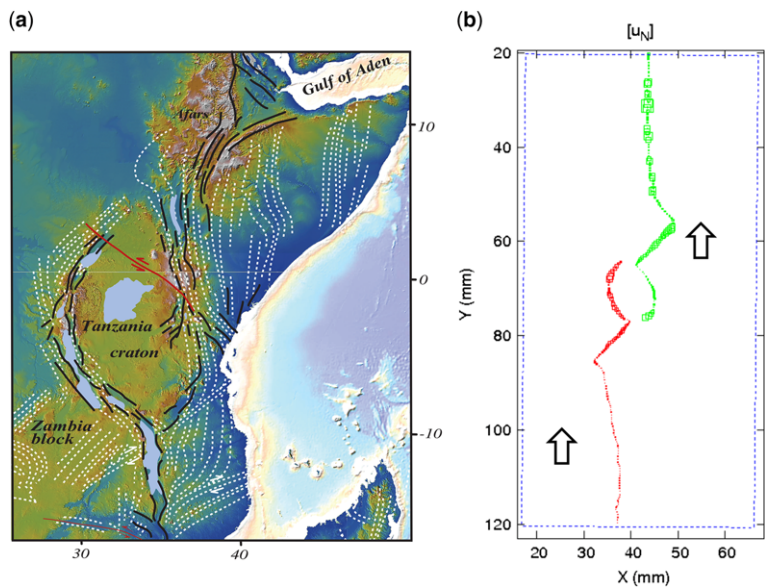


Fig. 2. The role of pre-existing weakness in plate rupture. (a) The correspondence between normal faults of the East Africa Rift (black lines) and the orientation of Proterozoic pre-existing sutures (white dashed lines). The red line shows a transpressional shear zone that cross-cuts the region (based on Tommasi & Vauchez 2015). (b) Rifting in a roughly 5×10 cm plate, 3.5 cm thick (adapted from Hall *et al.* 2006). The plate is subject to compression, as indicated by the black arrows. The short (1.2 cm) straight line segments in red and green and oriented in a 45° direction are sites of pre-existing weaknesses that have been scored in the plate. The red and green symbols show failure locations along the plate.

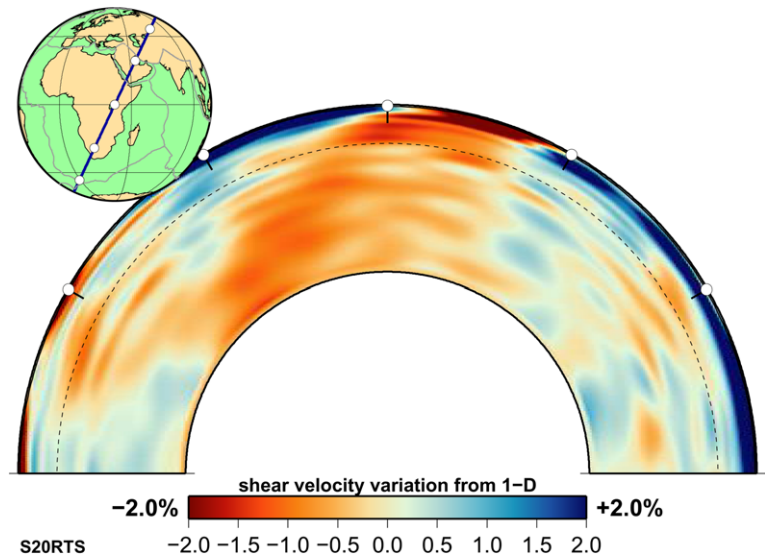


Fig. 3. Image of shear wave velocity structure beneath Africa in the tomographic model S20RTS (after Ritsema *et al.* 1999) along the profile indicated in the smaller map-view inset. The large low shear velocity province starts at the core–mantle boundary beneath southern Africa and trends upwards in a northeasterly direction towards the Red Sea. The colour scale is $\pm 2\%$ of a one-dimensional (1-D) shear velocity model.

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large (degree-two in spherical harmonics) features of planet Earth that extend upwards from the core–mantle boundary (the other lies beneath the south-central Pacific Ocean). These are sometimes called ‘super-plumes’ and are thought to be long-lived thermo-chemical anomalies that may spawn smaller plumes and may even be associated with the location of diamond formation (Torsvik *et al.* 2010). The ‘super-swells’ seen in the South Pacific (McNutt 1998) and African (Nyblade & Robinson 1994) regions are attributed to these anomalously buoyant regions of the mantle. We note that most studies of the dynamic topography of these features are interpreted in terms of thermal effects. The effects of compositional differences is a challenge as there is still no consensus about the chemical nature (if any) of these anomalies; see Davies *et al.* (2015) for a discussion of LLSVPs.

Current tomographic models of the seismic velocity structure of the mantle can be used to predict the time-dependent mantle flow using current images as boundary conditions for backwards advection (Bunge *et al.* 2002; Conrad & Gurnis 2003; Forte *et al.* 2010; Moucha & Forte 2011). Conrad & Gurnis (2003) modelled the evolution of dynamic topography beneath the region of Africa (Fig. 4). At 105 Ma, the most pronounced uplift

was located at the site of Gondwana rifting between what is now present day Africa and India. By 60 Ma, Africa had migrated southwards and the site of highest uplift was East Africa; during the last 30 Ma this uplift has moved from eastern to southern Africa. At face value, the sites with the highest predicted uplift correspond with rifting in Africa over the past 105 Ma, suggesting that the dynamic topography plays an important part in African rifting. We next consider the rifting forces associated with dynamic topography and GPE.

Radial tractions and gravitational potential energy

Stresses applied to a plate in a radial direction lead to a dynamic topography, the magnitude of which is the ratio of the radial stresses to the density contrast between the plate and air/water multiplied by the gravitational acceleration. The stronger the source of buoyancy, the larger the stress and the dynamic topography. The dynamic topography is little affected by the presence of low viscosity channels because it is the result of normal stresses that are effectively transmitted across rheological boundaries (Naliboff *et al.* 2012). Figure 5 shows

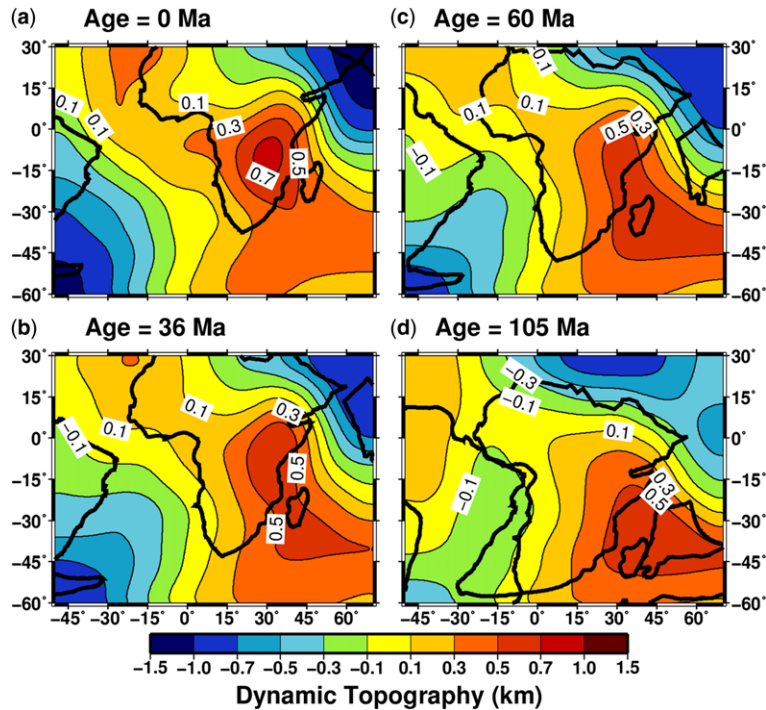


Fig. 4. Dynamic topography (in km) of Africa over the past 105 Ma (from Conrad & Gurnis 2003).

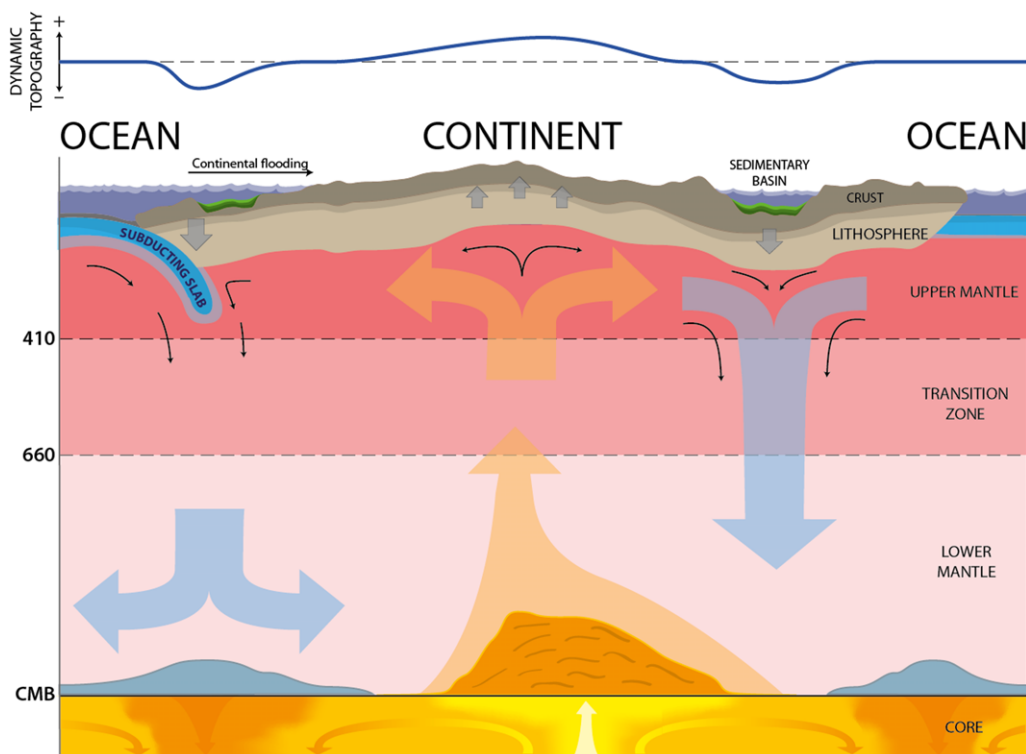


Fig. 5. Schematic diagram illustrating the sources of stress acting on the lithosphere. The sources include horizontal (thin black arrows) and radial tractions (thick orange and blue arrows) induced by flow in the mantle driven by density heterogeneities (i.e. slabs or large-scale upwellings) and gravity acting on intra-lithospheric density heterogeneities (the radial and lateral structure of the crust and lithosphere). The latter generates topography via isostasy. Radial tractions induce physical deflection of the lithosphere, giving rise to the dynamic topography.

the forces at play. Membrane stresses result from the extension and compression of the lithosphere as it bends and flexes. Gravitational stresses result from the topography and try to restore the plate to gravitational equilibrium. An example is the contributions to the ridge-push force that result from uplift of the ridge axis.

In addition to the gravitational stresses arising from the dynamic contribution to topography, we must consider those resulting from intra-lithospheric heterogeneity due to variations in the crustal thickness and density as well as those in the mantle lithosphere. No existing work has used a full lithospheric structure with variations in thickness and density and calculated the stresses because this is technically difficult. What is often calculated instead is the GPE arising from the density variations for equal thickness columns compensated at a typical depth, which are balanced by gradients in the deviatoric stress (Molnar & Lyon-Caen 1988). We can solve for the deviatoric stresses that balance the GPE (Flesch *et al.* 2001; Stamps *et al.* 2010;

Gosh & Holt 2012) or apply it as a pressure on all horizontal faces of a column (Lithgow-Bertelloni & Guynn 2004).

One complication in distinguishing the sub-lithospheric mantle v. lithospheric contributions to sources of stress arises when using the GPE formulation. Because this depends on the total topography, we must know which part of the topography is supported dynamically and which part is supported isostatically by variations in the crustal and lithospheric mantle thickness and density. This is often difficult or impossible to achieve because we do not know the lithospheric structure well enough and the dynamic contributions are derived from flow models. Nonetheless it is possible to attempt a separation by calculating a dynamic contribution from a given mantle density structure and subtracting it from the total topography and then calculating the GPE, as in Lithgow-Bertelloni & Guynn (2004).

In the context of Africa, recent work by Stamps *et al.* (2010) calculated the deviatoric stresses due to lateral gradients in the GPE for the region of

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Africa. Like Lithgow-Bertelloni & Guynn (2004), they used CRUST 2.0 in their models, although they did not separate dynamic from isostatic contributions. Based on models of lithospheric strength, they concluded that buoyancy forces are probably insufficient to rupture thick, cold continental lithosphere.

Figure 6 (coloured contours) shows the radial tractions that will give rise to the dynamic topography of the slab model and also two shear wave tomographic models: the Grand (Grand *et al.* 1997) and S20RTS (Ritsema *et al.* 1999) models. It is clear that the tomographic models do a much better job

of explaining the observed dynamic topography for Africa. This is not surprising because the slab model ignores buoyancy sources not related to subduction. In regions of long-lived subduction, the slab model is a more faithful representation of density heterogeneity and dynamic topography (e.g. the western Pacific) and is much better at predicting stresses than the World Stress Map (Lithgow-Bertelloni & Guynn 2004). It is also clear that differences between tomographic models and hence the mantle structure have a large influence on the azimuth and magnitude of the stresses that will eventually drive rifting. S20RTS predicts a broad area of

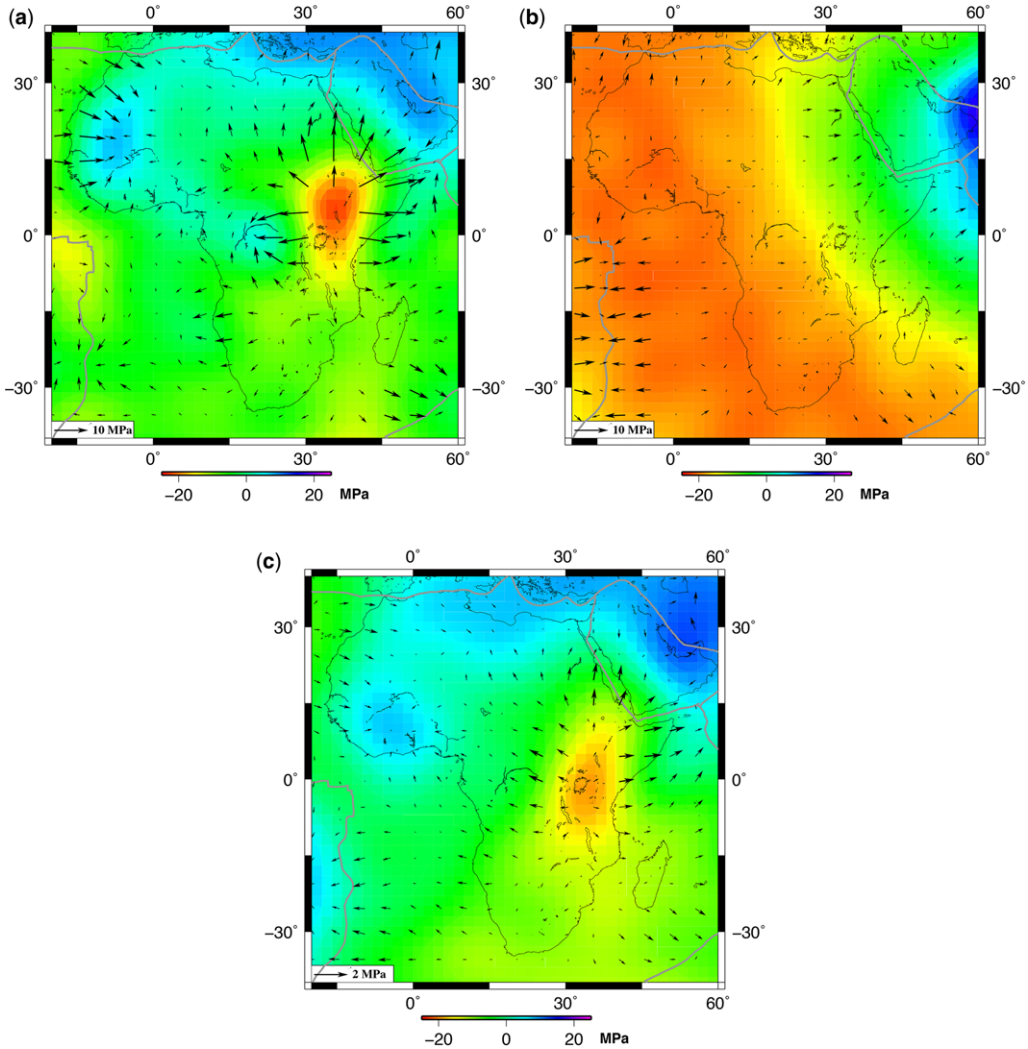


Fig. 6. Net horizontal tractions (arrows) and radial tractions (colours) for instantaneous mantle flow driven by density heterogeneities derived from (a) the shear wave tomographic model Grand (Grand *et al.* 1997), (b) the slab model (Lithgow-Bertelloni & Richards 1998) and (c) the S20RTS model (Ritsema *et al.* 1999).

outwards radial traction that roughly parallels the EAR and agrees with the observed dynamic topography, not only in eastern Africa, but also in northern Africa. The Grand model (Grand *et al.* 1997) does not do as good a job of predicting the extent of the dynamic topography – it is lacking in southern Africa and northern Africa. However, this model predicts larger extensional stresses in an east–west direction along the EAR (Fig. 7).

Figure 7 shows the predicted deviatoric stresses from the GPE (model TD0 in Lithgow-Bertelloni & Gynn 2004) and the net horizontal and radial tractions calculated from the flow induced by the density anomalies inferred from the Grand tomographic model. The magnitude of the GPE contribution is on the order of 25 MPa over the EAR (Fig. 6a). This magnitude depends on the details of the lithospheric structure used and the mechanism

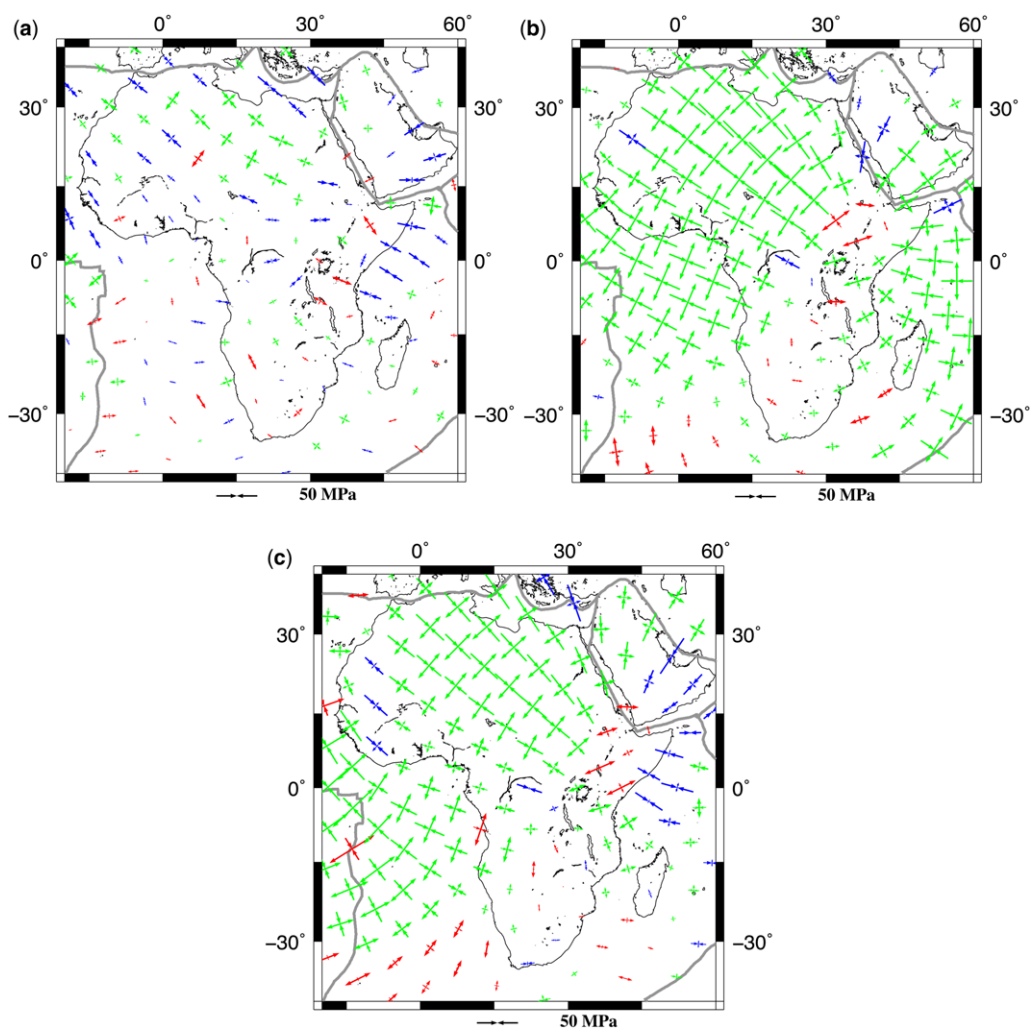


Fig. 7. Total predicted deviatoric stresses in the lithospheric plate. Red indicates extension, green transform and blue compression. Stresses resulting from (a) GPE use the lithospheric model TD0 from Lithgow-Bertelloni & Gynn (2004). Net (b) horizontal and (c) radial tractions are calculated from the flow induced by heterogeneity derived from the Grand tomographic model. The deviatoric stresses are calculated by applying radial and horizontal tractions to an elastic shell 100 km thick (see Lithgow-Bertelloni & Gynn 2004). Stresses due to horizontal tractions induce a largely strike-slip regime because, in the absence of radial tractions, the stress resultants are equal and opposite (Lithgow-Bertelloni & Gynn 2004). The small non-strike-slip regimes are not significant as they correspond to very small stresses.

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of isostatic compensation (Lithgow-Bertelloni & Guynn 2004; Naliboff *et al.* 2012). The most recent crustal compilation, CRUST 1.0 (Laske *et al.* 2013), leaves this magnitude largely unchanged. The stresses associated with radial tractions are larger (30–70 MPa) across the region. With both the GPE and radial tractions the sense of rifting in the EAR is correct (i.e. extension in a roughly east–west direction near the EAR). The stress magnitudes are large in west Africa, but in a strike-slip regime, and very small in southern Africa.

Horizontal tractions at the base of the African lithosphere

Viscous coupling between the mantle and the overlying lithosphere ensures that the horizontal shear tractions generated by mantle flow will act at the base of the plate (Fig. 5). These are the tractions that primarily drive all plate motions, regardless of the location of the buoyancy sources. However, buoyancy sources closer to the plate will provide a larger contribution to the total torque balance for that plate. For example, plates with slabs attached to them will be affected more strongly and directly by the slab buoyancy and strength than plates without slabs (Conrad & Lithgow-Bertelloni 2002). In Africa, the Somalian plate has no attached slab (although we note that Forte *et al.* (2010) argued that subduction may be initiating in this region) and the Nubian plate has minimal slab activity to the north. African plate dynamics have a more significant contribution from the buoyancy of the African LLSVPs (Lithgow-Bertelloni & Silver 1998).

Consider a high-density blob sinking into the mantle. This will pull mantle material towards the downwelling, which will impose tractions on the base of the plate and, depending on their relative directions of movement, can both enhance or oppose plate motion. In contrast, resisting tractions are those that result from the plate moving over a viscous mantle (perhaps in response to distant forces) and will always oppose plate motion. The net traction is the sum of the driving and resisting tractions at a given point and is the net traction that drives deformation (i.e. contributes to rifting).

How effectively the horizontal tractions drive deformation will depend on the presence or absence of low viscosity channels in the mantle and the lithosphere.

Figure 6 (black arrows) shows the predicted horizontal tractions from the Grand, S20RTS and slab models. Figure 7b shows the stresses induced by these tractions, but only for the Grand model. We note that allowing for reasonable lateral variations in viscosity has little effect on the results (Naliboff *et al.* 2009).

The maximum magnitudes are 30–60 MPa and the sense of stress is conducive to rifting in the northern parts of the EAR (i.e. extensional). Note that there is some extension in southern Africa, but the stress magnitudes are negligible. These results are in rough agreement with those of Forte *et al.* (2010) and Moucha & Forte (2011), although the approach taken is rather different from that of Lithgow-Bertelloni & Guynn (2004).

Combined effects: a full plate model with radial and horizontal traction

Figure 8 shows the predicted stresses when combining all sources of stress (GPE, horizontal and radial

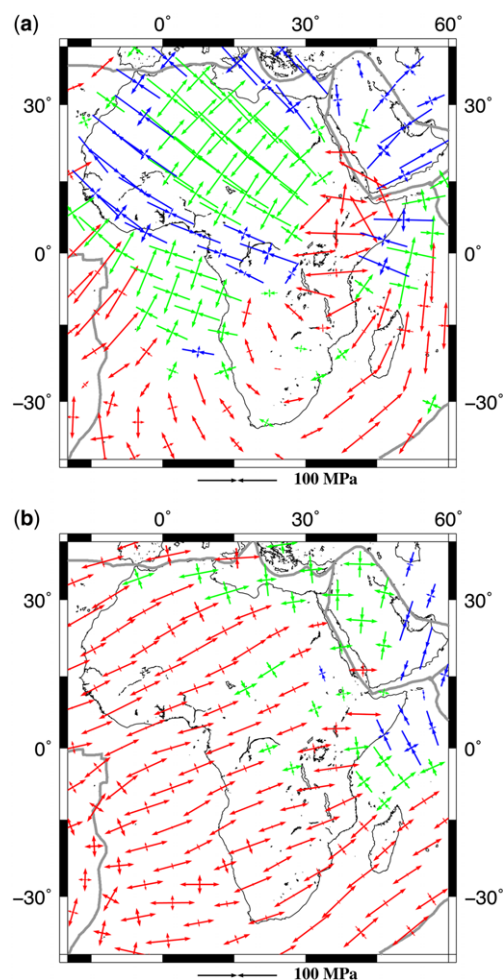


Fig. 8. Stresses obtained when summing all contributions to the stress field (mantle + GPE) for (a) the Grand and (b) slab models. See Figure 7 for more detail.

mantle tractions) for two mantle heterogeneity models (the Grand and slab models).

Figure 8b shows the predicted radial and horizontal stress for a model driven solely by slabs descending into the mantle. As noted, this model works well at predicting stresses in areas of previous long-lived subduction, such as SE Asia (Lithgow-Bertelloni & Gynn 2004). However, it does not do a good job of fitting Africa's dynamic topography and it predicts very small horizontal tractions (<10 MPa) and elastic stresses (<25 MPa) across most of the African continent. This is not surprising because the slab model has no buoyancy associated with active upwelling (i.e. no LLSVPs). It is clear that the mantle structure beneath Africa and its dynamic consequences plays a fundamental part in dynamic uplift and rifting in East Africa.

The Grand model (Fig. 8a) predicts much larger stresses and the right sense of extension in the EAR. An inventory of forces yields a GPE contribution to the stress on the order of 25 MPa and similar magnitudes, perhaps as high as 70 MPa, from horizontal and radial tractions (i.e. the dynamic topography). The total sum is no more than 100 MPa, well short of the *c.* 0.5–1.0 GPa needed to break thick, cold continental lithosphere.

Geophysical signature of lithospheric melt

The EAR has been the focus of a number of geophysical investigations of the rift system for many decades – for example, the early refraction work in Ethiopia (Berckhemer *et al.* 1975) and the KRISP experiment in Kenya (Khan *et al.* 1999). The Ethiopia Afar Geoscientific Lithospheric Experiment (Maguire *et al.* 2003) and the Ethiopia Kenya Broadband Seismic Experiment (Nyblade & Langston 2002) marked the start of a series of deployments of broadband seismometers across the Horn of Africa. Over 200 seismic stations have been installed in the region in the past 15 years; see, for example, the compilations in Gallacher *et al.* (2016, in review) and Civiero *et al.* (2015). In contrast, the coverage prior to this period was based on a handful of seismic stations. A drive to better instrumentation in Ethiopia has come from leadership at the University of Addis Ababa (e.g. Ayele *et al.* 2007). Needless to say, our 'picture' of the crust and underlying mantle has seen a stepwise improvement. For example, as station numbers have increased, body-wave tomographic images of the velocity structure have steadily improved in both coverage and resolution (e.g. Bastow *et al.* 2005, 2008; Benoit *et al.* 2006; Hammond *et al.* 2013; Civiero *et al.* 2015).

The 2005 Dabbahu event provided a particularly spectacular exhibition of dyke injection along a

rift segment (e.g. Wright *et al.* 2006; Ayele *et al.* 2007; Ebinger *et al.* 2008; Hamling *et al.* 2009; Grandin *et al.* 2010), which was the first event of its kind to be documented with modern geodetic techniques such as InSAR. The Afar Rift Consortium project has led to a concurrent multidisciplinary programme of study, such that we now have a much better understanding of how the crust and mantle are deforming during rifting (Wright *et al.* 2016).

The seismic signature of rifting reveals a number of interesting features. Figure 9 shows an image of the P- and S-wave velocity structure at a depth of 80 km (Hammond *et al.* 2013), near the bottom of the African lithosphere in this region. The velocities are not absolute, but are derived using relative travel-time tomography where the velocity anomalies are perturbations from a background mean. The rift is marked by punctuated upwellings of hot and partially molten material that penetrate the plate. These low velocity zones are not confined to areas beneath magmatic segments in the rift valley; many of these zones are distributed along the margins of the rift. An example is the Silty–Debra–Zeit lineament on the western edge of the Main Ethiopian Rift and the western shoulder of the rift bordering on Afar (near Mekele). The low velocity anomalies show a remarkable resemblance to the overlying geometry of the rift, even exhibiting sharp right-angle bends away from the trend of the Main Ethiopian Rift to the trend of the Red Sea rift segments. In general, the Afar region shows complicated variations in upper mantle velocity structure (e.g. Stork *et al.* (2013), who used Pn tomography to image the uppermost mantle). The low velocities cannot be explained by thermal anomalies and partial melt is required (see Bastow *et al.* 2005). It is worth noting that the absolute travel-time delays in this region are the largest anywhere on Earth.

Surface-wave tomography reveals similar features and offers better vertical resolution, but poorer lateral resolution (Fishwick 2010). Using teleseismic Love and Rayleigh waves, Sebai *et al.* (2006) and Sicilia *et al.* (2008) reported low velocities to depths of roughly 400 km beneath the Main Ethiopian Rift, the Afar hotspot, the Red Sea, the Gulf of Aden and east of the Tanzania Craton. They saw a stratified anisotropic structure, but noted that their lateral resolution was of the order of 500 km. Guidarelli *et al.* (2011) used regional Rayleigh waves to image the Afar region in finer detail, revealing a patchwork of stranded continental slivers surrounded by regions of low velocity, melt-rich material, a pattern not unlike that reported by Stork *et al.* (2013). Using ambient noise imaging, Korostev *et al.* (2015) showed how magmatic processes are focused at the axial magmatic segments and also

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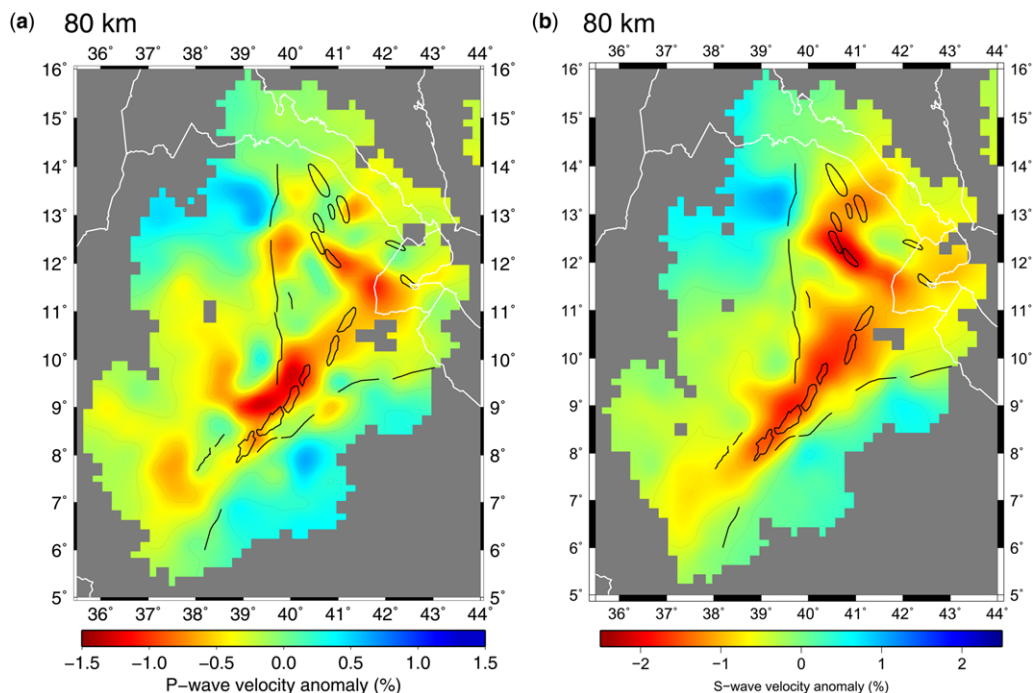


Fig. 9. (a) P-wave and (b) S-wave velocity structure of the mantle at a depth of 80 km. Velocities are given as a percentage difference from the one-dimensional background mean (grey indicates regions of unresolved velocities). Black lines mark the rift border faults and the black elliptical shapes mark magmatic segments. Figure is from the model of Hammond *et al.* (2013).

at the rift flanks. Gallacher *et al.* (2016, in review), using array-based Rayleigh wave tomography with teleseismic phases, found heterogeneities in the Main Ethiopian Rift as well as the Afar region. They also imaged the anisotropic velocity structure of the region to a depth of 250 km. Their contrasting observations in the Main Ethiopian Rift and Afar region suggest that melt production is at its highest early during the break-up process. In agreement with other studies, they noted a pronounced segmentation in the low velocity zones that extended well into the mantle.

The velocity structure of a plate shows discontinuities at the base of the crust (the Moho) and, at times, even at the base of the plate (the lithosphere–asthenosphere boundary). The so-called receiver functions highlight these discontinuities through the identification of seismic phase conversions (e.g. S- to P-wave or P- to S-wave at these boundaries; e.g. Dugda *et al.* 2005). Using P- to S-wave conversions and their reverberations, Stuart *et al.* (2006) and Hammond *et al.* (2014) presented a comprehensive picture of the crustal structure across most of Ethiopia and Eritrea (Fig. 10a). Figure 10b shows a transect from the Ethiopian plateau to northern Eritrea (at a latitude of roughly 13°).

The crust thins from nearly 40 km in thickness to <15 km in a dramatically short distance. Large variations in the crustal structure of Afar were also noted in the early seismic refraction and gravity work of Makris & Ginzburg (1987). More recently, Reed *et al.* (2014) showed dramatic variations in the crust in a transect across the Tendaho Graben in the Afar region. The thinnest crust is marked by an increased the V_p/V_s ratio, widely well above 2; such values are almost impossible to explain without the presence of fluid (partial melt in this instance). Hammond (2014) has shown that anisotropy due to melt alignment, both horizontally (sills) and vertically (dykes), is required to explain the crustal receiver functions in the Afar region.

The lithosphere–asthenosphere boundary in this region also exhibits rapid lateral changes in properties. Rychert *et al.* (2012) isolated S- to P-wave conversions at a depth of *c.* 80 km. Beneath the Ethiopian plateau, the conversions are consistent with a sharp transition from low velocity asthenosphere to the higher velocity cold and rigid pan-African lithosphere. The conversion is opposite in nature beneath the rift and suggests a transition to a lower velocity mantle above 80 km, an observation that Rychert *et al.* (2012) interpreted as the

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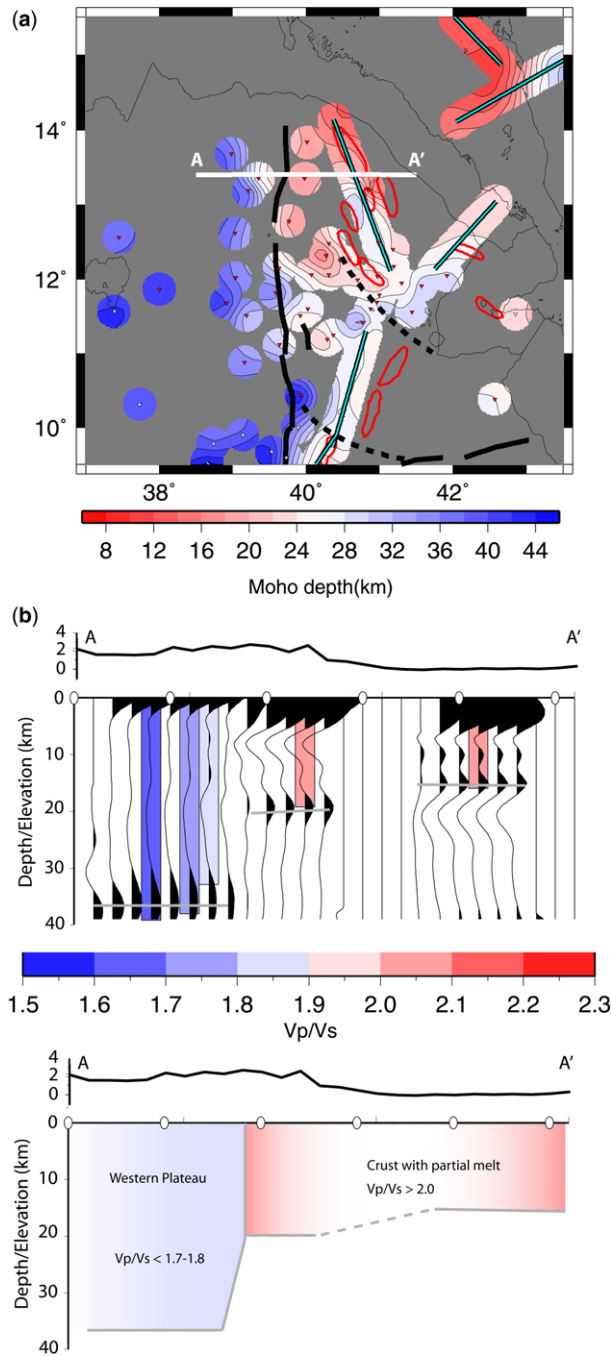


Fig. 10. The crustal structure of the Horn of Africa (from Hammond *et al.* 2011). **(a)** Map view of crustal thickness (dark grey areas are unresolved). Red ellipsoidal shapes mark the magmatic segments and the inverted triangles show the stations contributing to the crustal thickness estimates. The black lines show the border faults separating Afar from the western and southeastern plateaus. The dashed line shows the Tendahao-Gob'a discontinuity (northern line) and arcuate accommodation zone (southern line). **(b)** A west-east cross-sectional image of crustal structure along the latitude of *c.* 13° N produced using common conversion point imaging (see Hammond *et al.* (2011) for more detail). The lower part shows an interpretation with an estimate of average V_p/V_s ratios. $V_p/V_s > 2.0$.

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onset of melting in the rifted region. Again, the transition from the pan-African lithosphere to the rifted region is very sharp at the base of the plate. Based on a self-consistent model of mantle flow, Armitage *et al.* (2015) argued that the warmed lithosphere is thinned to <50 km beneath the rift, but the origin of the melt-related discontinuity at deeper depths is still unclear. Both the Moho and the lithosphere–asthenosphere boundary observations show that strain is localized at the rift flanks in a narrow region and do not support ideas of a broad, distributed stretching of the plate, as suggested by the tectonic stretching model.

Observations of seismic anisotropy also highlight regions of intense melt alignment, both at shallow depths beneath the magmatic segments (Keir *et al.* 2005, 2011) and throughout the lithosphere beneath the rift flanks (Kendall *et al.* 2005, 2006; Hammond & Kendall 2016). Finite frequency modelling of SKS splitting further supports this conclusion (Hammond *et al.* 2010). Bastow *et al.* (2010) used a combined analysis of SKS splitting and azimuthal variations in Love and Rayleigh waves to show that the anisotropy in the upper 80 km of the Main Ethiopian Rift must be due to aligned melt. The degree of alignment is particularly intense in the narrowest regions of the Main Ethiopian Rift (Hammond & Kendall (2016).

However, studies based on SKS splitting (Gashawbeza *et al.* 2004; Kendall *et al.* 2006; Gao *et al.* 2010) and joint SKS splitting and receiver function analysis (Obrebski *et al.* 2010) have argued that there must be a deeper seated layer of anisotropy reflecting the current asthenospheric flow. Using shear wave splitting tomography, Hammond *et al.* (2014) showed that SKS splitting across the region can indeed be explained by multiple causes. The sub-lithospheric mantle in the rift region shows a consistent pattern of anisotropy across the region that can be explained by the lattice-preferred orientation of olivine due to mantle flow. The anisotropy and isotropic travel-time tomography delineate a carpet of low velocity material that is rising upwards from the deeper mantle, migrating in a northeasterly direction towards the Arabian craton. The lithospheric anisotropy shows much more variability. In some places it is small and can be explained by pre-existing fabrics in slivers of continental crust, whereas in other areas it is large and aligned with the low velocity anomalies and/or magmatic segments (as noted also by Keir *et al.* 2005, 2011).

Magnetotelluric experiments provide complementary geophysical methods to image the presence of melt. Whaler & Hautot (2006) showed evidence of crustal melt beneath magmatic segments of the Main Ethiopian Rift and even in the lower crust beneath the flank of the Ethiopian

plateau. There appears to be large volumes of melt beneath parts of Afar that extend well into the mantle (Desissa *et al.* 2013). There is some discrepancy between the estimated melt volumes from magnetotelluric and seismic methods, but the effect of melt composition may explain some of this (Pommier & Garnero 2014). Linking seismic and electrical images of melt and their anisotropy is an ongoing challenge.

Cumulatively, these observations suggest that strain is localized along narrow zones and that melt is localized on the marginal lithosphere–asthenosphere boundary (see Holtzman & Kendall 2010 for a more extensive discussion). There is no evidence of a broad zone of stretching, implying that local processes play a central part in rifting.

Discussion and conclusions

We have calculated the stresses induced by mantle flow, crustal structure and topography in two types of models using a finite element model of the lithosphere. In one model, the mantle flow is derived from a model exclusively driven by slab subduction (the slab model), whereas in the other model it is derived from a shear wave tomographic model (the Grand model). Based on an inventory of the stresses available, the Grand model predicts extensional stresses of up to 100 MPa. In contrast, the slab model predicts stresses of no more than 25 MPa. The maximum magnitudes can vary between tomographic models as they depend on the amplitude of the seismic anomalies. The conversion of velocity to density could potentially increase the magnitudes with these models by a maximum factor of two. It is difficult to decouple the contribution from various mechanisms, but the mantle structure beneath Africa plays a key part in providing the radial (and hence dynamic topography) and horizontal tractions in addition to the GPE necessary for rifting. Distant slab forces are clearly not enough to explain rifting in this region.

Nevertheless, the combined stresses are clearly not sufficient to rupture thick, cold, dry continental lithosphere (>1 GPa). There are at least three reasons why the yield strength might be lower than that required in such a model. First, the presence of melt weakens the lithosphere and, based on the calculations of Buck (2004), stresses of *c.* 200 MPa are needed, which are arguably still more than those available. However, a higher geothermal gradient will reduce this figure; more work is required to test the sensitivity of the yield stress to this and other factors (e.g. melt composition and fluid content). Second, pre-existing weaknesses and lithospheric thin spots will reduce the strength of the plate. There is considerable evidence

worldwide that plates rupture along pre-existing sutures (Wilson 1966) and Africa is no exception (Vauchez *et al.* 1997). Third, the mantle may be much wetter than we think it is and the 'jelly sandwich' model of continental lithosphere may overestimate the yield strength of the mantle (Jackson 2002). However, these models still predict a very strong crust. In summary, based on our current knowledge of the strength of the continental lithosphere, it is increasingly clear that we must invoke the presence of melt as a mechanism for reducing the strength of the African plate, in addition to pre-existing weaknesses and/or lithospheric thin spots.

Recent geophysical investigations show abundant evidence for strain localization and focused melt upwellings beneath East Africa. These observations provide the 'smoking gun' for magma-assisted models of rifting and do not support the tectonic stretching models. A better estimate of the variations in plate thickness across the region would provide useful information to give a better understanding of plate–mantle interactions (e.g. Ebinger & Sleep 1998).

There are still a number of outstanding issues and poorly known parameters in understanding continental rifting. For example, do we really know the strength of a plate? The yield stresses are better known for the crustal component of the plate, but there is some uncertainty in determining the strength of the mantle component. If all sources of density heterogeneity in the mantle and lithosphere were known, along with all the rheological properties, it would be possible to model mantle flow and surface deformation within the same framework. However, there would still be technical difficulties because of the vastly different time-scales of the processes involved in different types of deformation from elastic (instantaneous) to viscous (millions of years), which are at present impossible to model in the context of one code.

Another uncertainty is the style of mantle upwelling beneath Africa. The so-called Afar plume is not a simple diapiric plume. The architecture of mantle upwelling from the core–mantle boundary to the surface beneath NE Africa is still being determined and there is considerable debate as to its thermo-chemical nature. Smaller plumes may initiate from instabilities in a deeper super-plume that may or may not be chemically distinct from the ambient mantle (Civiero *et al.* 2015; Thompson *et al.* 2015). However, it is hard to see how this anomaly could not be buoyant given the significant dynamic topography of Africa.

The analysis in this work concentrates on Africa and the EAR system. However, what about other continental rifts? Although subduction is not a significant driver of rifting in East Africa, it plays a more significant part in rifting in the Red Sea

(Bellahsen *et al.* 2003). An understanding of rifting in local settings must consider the processes as a coupled system on a global scale.

The work presented here resulted from discussions at the Magmatic Rifting and Volcanism Conference held in Addis Ababa in 2012 and we would like to thank the organizers of the conference for inviting CLB to attend. We thank James Hammond and Chris Hawkesworth for helpful suggestions, which have improved the writing. A number of people assisted in the production of the figures and their help is gratefully appreciated: James Wookey for Figures 3 and 5, Steve Hall and Alain Vauchez for parts of Figure 2, Clint Conrad for Figure 4, James Hammond for Figures 9 and 10 and John Naliboff for help with Figures 7 and 8. We are grateful to reviewer Roger Buck and editor Tim Wright for their constructive comments and patience. This work was supported by NERC grant NE/E007414/1.

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